# REPORT DOCUMENTATION PAGE

| 1. REPORT DATE (DD-MM-YYYY) | 2. REPORT TYPE | 3. DATES COVERED (From - To) |
|---|---|---|
| 23-01-2008 | Journal Article | |

| 4. TITLE AND SUBTITLE | 5a. CONTRACT NUMBER |
|---|---|
| Wind Stress Drag Coefficient Over the Global Ocean | |
| | 5b. GRANT NUMBER |
| | 5c. PROGRAM ELEMENT NUMBER |
| | 0602435N |

| 6. AUTHOR(S) | 5d. PROJECT NUMBER |
|---|---|
| Ahmet Birol Kara, Alan J. Wallcraft, E. Joseph Metzger, Harley E. Hurlburt, Chris Fairall | |
| | 5e. TASK NUMBER |
| | 5f. WORK UNIT NUMBER |
| | 73-8677-06-5 |

| 7. PERFORMING ORGANIZATION NAME(S) AND ADDRESS(ES) | 8. PERFORMING ORGANIZATION REPORT NUMBER |
|---|---|
| Naval Research Laboratory<br>Oceanography Division<br>Stennis Space Center, MS 39529-5004 | NRL/JA/7320-06-7059 |

| 9. SPONSORING/MONITORING AGENCY NAME(S) AND ADDRESS(ES) | 10. SPONSOR/MONITOR'S ACRONYM(S) |
|---|---|
| Office of Naval Research<br>800 N. Quincy St.<br>Arlington, VA 22217-5660 | ONR |
| | 11. SPONSOR/MONITOR'S REPORT NUMBER(S) |

**12. DISTRIBUTION/AVAILABILITY STATEMENT**

Approved for public release, distribution is unlimited.

**13. SUPPLEMENTARY NOTES**

**14. ABSTRACT**

Interannual and climatological variations of wind stress drag coefficient are examined over the global ocean from 1998 to 2004. Here cD is calculated using high temporal resolution (3- and 6-hourly) surface atmospheric variables from two datasets: 1)the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and 2) the Navy Operational Global Atmospheric Prediction System. The stability-dependent cD algorithm applied to both datasets gives almost identical values over most of the global ocean, confirming the validity of results. Overall, major findings of this paper are as follows: 1) the cD value can change significantly on 12hourly time scales around the Kuroshio and Gulf Stream current systems; 2) there is strong seasonal variability in cD, but there is not much interannual change in the spatial variability for a given month; 3) a global mean Cd= 1.25 * 10-³ is found in all months, while cD is less than or equal to 1.0 * 10-³ is prevalent over the North Pacific and North Atlantic Oceans and in southern high-latitude regions as well, and cD is less than or equal to 1.0 * 10-³ is typical in the eastern equatorial Pacific cold tongue; and 4) including the effects of air-sea stability on cD generally causes an increase of less than 20% in comparison to the one calculated based on neutral conditions in the tropical regions. Finally, spatially and temporally varying cD fields are therefore needed for a variety of climate and air-sea interaction studies.

**15. SUBJECT TERMS**

global ocean, NOGAPS, ECMWF, drag coefficient

| 16. SECURITY CLASSIFICATION OF: | | | 17. LIMITATION OF ABSTRACT | 18. NUMBER OF PAGES | 19a. NAME OF RESPONSIBLE PERSON |
|---|---|---|---|---|---|
| a. REPORT | b. ABSTRACT | c. THIS PAGE | | | Ahmet Birol Kara |
| Unclassified | Unclassified | Unclassified | UL | 9 | 19b. TELEPHONE NUMBER (Include area code)<br>228-688-5437 |

# NOTES AND CORRESPONDENCE

## Wind Stress Drag Coefficient over the Global Ocean*

A. BIROL KARA, ALAN J. WALLCRAFT, E. JOSEPH METZGER, AND HARLEY E. HURLBURT

*Oceanography Division, Naval Research Laboratory, Stennis Space Center, Mississippi*

CHRIS W. FAIRALL

*Physical Science Division, NOAA/Earth System Research Laboratory, Boulder, Colorado*

### ABSTRACT

Interannual and climatological variations of wind stress drag coefficient ($C_D$) are examined over the global ocean from 1998 to 2004. Here $C_D$ is calculated using high temporal resolution (3- and 6-hourly) surface atmospheric variables from two datasets: 1) the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and 2) the Navy Operational Global Atmospheric Prediction System (NOGAPS). The stability-dependent $C_D$ algorithm applied to both datasets gives almost identical values over most of the global ocean, confirming the validity of results. Overall, major findings of this paper are as follows: 1) the $C_D$ value can change significantly (e.g., >50%) on 12-hourly time scales around the Kuroshio and Gulf Stream current systems; 2) there is strong seasonal variability in $C_D$, but there is not much interannual change in the spatial variability for a given month; 3) a global mean $C_D \approx 1.25 \times 10^{-3}$ is found in all months, while $C_D \geq 1.5 \times 10^{-3}$ is prevalent over the North Pacific and North Atlantic Oceans and in southern high-latitude regions as well, and $C_D \leq 1.0 \times 10^{-3}$ is typical in the eastern equatorial Pacific cold tongue; and 4) including the effects of air–sea stability on $C_D$ generally causes an increase of >20% in comparison to the one calculated based on neutral conditions in the tropical regions. Finally, spatially and temporally varying $C_D$ fields are therefore needed for a variety of climate and air–sea interaction studies.

## 1. Introduction and motivation

The momentum exchange between the atmosphere and ocean through wind stress is of importance for many purposes, including air–sea interaction studies, climate studies, ocean modeling, and ocean prediction. Wind stress is typically obtained from bulk parameterizations that estimate turbulent fluxes using standard meteorological data (e.g., Fairall et al. 2003). In particular, the total wind stress magnitude ($\tau$) at the ocean surface can be calculated from the square of the wind speed at 10 m above the sea surface ($V_a$), the density of air ($\rho_a$), and a dimensionless drag coefficient ($C_D$) with $\tau = \rho_a C_D V_a^2$.

Previously, attention has been given to construction of wind stress climatologies over the global ocean (Chelton et al. 1990) and experimental analysis of $C_D$ at a few particular locations (Donelan et al. 1997). Based on the authors' knowledge there is no quantitative study examining the spatial and temporal variability of $C_D$, a parameter that is used for calculating $\tau$, over the global ocean. Experimental measurements for $C_D$ are rarely available, and those that are available do not have sufficient temporal and spatial resolution to determine its global distribution. In addition, prior to the 1990s there were uncertainties in the accuracy of near-surface meteorological variables from the existing archived numerical weather prediction centers, such as the European Centre for Medium-Range Weather Forecasts (ECMWF) and the National Centers for Environmental Prediction (NCEP), preventing the accu-

rate calculation of $C_D$ based on bulk variables. For example, a study that explores climatological mean features of $C_D$ over the global ocean was presented by Trenberth et al. (1989). That study was limited to calculating near-surface stability based on monthly mean atmospheric variables ($2.5° \times 2.5°$ resolution) at the 1000-mb level from ECMWF (1980–86), which was the only available product from ECMWF at that time. That study did not examine diurnal and interannual variations in $C_D$ either.

The quality of archived products, such as these from ECMWF, has greatly improved since that time. The amount of observational data has greatly increased and better assimilation methods have been developed. Thus, it is now possible to better determine air–sea stability and to investigate spatial and temporal variations (diurnal, interannual, and climatological) of $C_D$ over the global ocean—the major goal of this paper. This will be accomplished using an algorithm, which fully takes the air–sea stability into account.

A study that examines the global variability of $C_D$ is desirable given the fact that the use of an inaccurate $C_D$ in calculating wind stress may result in serious errors. As an example, ignoring the effects of water vapor flux in the parameterization of $C_D$ can give a wind stress value that is $\approx 6$ times less than its actual value at very low wind speeds (Kara et al. 2005). This is particularly true on short time scales because of air–sea stability (e.g., day and night). In other words, averaging standard meteorological variables over a day may generally give neutral conditions (e.g., the difference between near-surface air and sea surface temperature is almost zero and relative humidity is 100%). This is generally not the case for daytime versus nighttime conditions. For this reason, $C_D$ needs to be calculated at fine temporal resolution (e.g., every 3 or 6 h) and then averaged over a day.

Given the need for proper determination of stability-dependent $C_D$ in calculating the wind stress magnitude over the global ocean, the main focus of this paper is threefold: 1) present spatial and temporal variations in $C_D$ for use in climate studies, 2) determine regions where there is strong/weak seasonal variability, and 3) reveal if there is any interannual variability in $C_D$ or if substituting a climatological mean may be appropriate over the different regions of the global ocean.

## 2. Methods and data

The parameterization of $C_D$ is still an active field with considerable diversity of approaches available in the literature (e.g., Taylor and Yelland 2001; Hwang 2005). Simply stated, the drag coefficient is usually rep-

resented as a combination of a wind speed–dependent neutral coefficient (or, equivalently, a roughness length), a hydrostatic stability dependence, and in some cases an enhancement at low wind speeds associated with wind gustiness (e.g., Fairall et al. 2003).

Various formulations of $C_D$ via surface roughness are available. They are based on friction velocity (rather than $V_a$) and do not include any air–sea stability dependence on $C_D$ using air–sea temperature difference or vapor mixing ratio values. There are also $C_D$ parameterizations based solely on $V_a$ (e.g., Trenberth et al. 1989). A constant $C_D$ value has been used in many studies (e.g., Kessler and Kleeman 2000; Sura et al. 2000; Koracin et al. 2004). Obviously, wind stresses that are calculated using such constant values exclude the significant changes in magnitude that can occur due to effects of air–sea stability in $C_D$. Several formulations include a simple form for stability effects on $C_D$ based on air–sea temperature difference (e.g., Smith 1988). However, water vapor flux is also an important parameter determining stability, especially in tropical regions (Kara et al. 2005). In particular, because of the exponential increase of saturation vapor pressure with temperature, in some regions, humidity has a first-order effect on the stability.

For the reasons mentioned above, a parameterization that takes full account of stability in calculating $C_D$ is required. The detailed parameterization presented in Kara et al. (2005) is used here. The formulation is based on the state-of-the-art Coupled Ocean–Atmosphere Response Experiment (COARE) bulk algorithm (version 3.0), which employs a turbulence theory based on the iterative estimations of the scaling variables to determine stability-dependent $C_D$. The stability-dependent $C_D$, as used in this paper, is expressed as simple polynomial functions of air–sea temperature difference, using air temperature at 10 m, $V_a$ at 10 m, and relative humidity at the air–sea interface to parameterize air–sea stability. Because of deficiencies in the COARE algorithm itself at high winds and ongoing debate, $C_D$ is kept constant in the parameterization for winds $>20$ m s$^{-1}$.

All variables for calculating $C_D$ over the global ocean are obtained from an archived gridded ($1° \times 1°$) product—the Fleet Numerical Meteorology and Oceanography Center (FNMOC) Navy Operational Global Atmospheric Prediction System (NOGAPS; Rosmond et al. 2002). NOGAPS is particularly chosen because it provides the above-mentioned atmospheric variables at high temporal frequency (3-hourly), a critical requirement to take air–sea stability into account. On the other hand, in section 4 we will also use similar 6-hourly atmospheric data from ECMWF to further confirm the

validity of $C_D$ values over the global ocean. The drag $C_D$ is computed from 1998 to 2004, the period when suitable archived NOGAPS data are available as of this writing.

Our availability of 3-hourly surface atmospheric variables from NOGAPS started in 2001; thus we use 6-hourly data from 1998 to 2000 and 3-hourly data from 2001 to 2004. The $C_D$ is first calculated at each 3 (or 6)-hourly time interval at each grid point over the global ocean. Monthly means are then formed. A time series of 1998–2004 may not be long enough to derive a climatological mean or to show interannual variability; however, it would still reveal general features of $C_D$ over the global ocean. An examination of $C_D$ performed over a long time period (e.g., 1979–2002) can be accomplished utilizing atmospheric variables from the ECMWF Re-Analysis, a topic which is mentioned in section 4.

## 3. Daily cycle of stability-dependent drag coefficient

The effect of stability on $C_D$ can be neglected on longer time scales (i.e., monthly). This is justified by the fact that the atmosphere is generally neutral or slightly unstable over the open ocean. Such conditions are only true for wind speed values between 6 and 25 m s$^{-1}$, as discussed in Bonekamp et al. (2002) in detail. However, wind speed can be outside this range over many regions of the global ocean. In addition, changes in stability through air–sea temperature and relative humidity differences near the sea surface do indeed exist when considering the diurnal cycle of near-surface atmospheric variables. This would certainly affect $C_D$, a stability-dependent parameter.

Thus, a few particular questions arise here: how variable (spatially and temporally) is $C_D$ for a given day? To answer these questions $C_D$ values calculated from the formulation of Kara et al. (2005) using data from NOGAPS (see section 2) are analyzed on 6 January 2004. This day is chosen just for illustrative purposes. Figure 1a clearly demonstrates large spatial variability in $C_D$ over the global ocean.

To examine the temporal variability of $C_D$ a few zoom regions, surrounding two major oceanic current systems (Gulf Stream and Kuroshio) and the western equatorial Pacific warm pool, are selected. The $C_D$ is shown at 0000Z and 1200Z (Fig. 1b). The increase or decrease in $C_D$ from 0000Z to 1200Z can be larger than 50% near regions around the Gulf Stream and Kuroshio as evident from the ratio values. As expected, this variability is due to the wind speed and air–sea stability

changing from 0000Z to 1200Z. For example, $V_a$ is 9.1 (11.9 m s$^{-1}$), air–sea temperature difference is 2.1° (−5.3°C), and relative humidity is 89% (66%) at 38°N, 70°W in the Gulf Stream region at 0000Z (1200Z). The corresponding $C_D \times 10^3$ value is 1.14 (1.56). The $C_D$ in the western equatorial Pacific has much less variability than that in the Gulf Stream and Kuroshio current systems.

Time periods (0000Z and 1200Z) were chosen randomly in order to investigate changes over a 12-h time period. The ratio of $C_D$ [i.e., $C_D$(at 0000Z)/$C_D$(at 1200Z)] is roughly out of phase in both the Gulf Stream and Kuroshio regions if one considers the day/night cycle (i.e., being day/night in the Gulf Stream and night/day in the Kuroshio). To capture the full diurnal change, the comparison in each region is also made between the $C_D$ map at the 3-hourly sampling time closest to high temperature and the map at the time closest to low temperature. The resulting ratio values clearly reveal that there are indeed large changes in $C_D$ from day to night in both regions (not shown), and the resulting ratios for $C_D$ generally resemble those shown in Fig. 1b.

Changes in $C_D$ mentioned in the preceding analysis were for a particular day. Thus, a lot of the differences could be temporal but not necessarily diurnal. Thus, we also investigated the mean diurnal variability of $C_D$ by averaging values at each 3-hourly interval from 0000Z to 1200Z and from 1200Z to 2400Z over a month. The ratio of $C_D$ during the two time periods is generally found to be within ±5% (±10%) during the Northern Hemisphere winter (summer).

Regarding the preceding analysis, we need to emphasize that the current parameterization of $C_D$ ignored the effects of ocean currents and waves. Such factors are particularly important, given that this paper focuses on several regions experiencing relatively strong currents, such as the Gulf Stream and the Kuroshio. For example, it is well known that especially in regions of strong currents, it is not simply the wind speed that is important but the vector difference in wind speed and ocean current speed and waves. In addition, possible impacts of ocean currents and wind waves on $C_D$ have been discussed in various studies (e.g., Wuest and Lorke 2003). On the other hand, a previous study based on a high temporal resolution global dataset demonstrated that strong ocean currents near the western boundaries (Kuroshio and Gulf Stream) do not substantially influence $C_D$ (Kara et al. 2007). This is due to the fact that the winds and currents are generally not aligned or locally correlated. In particular, that study revealed that the combined outcome of ocean currents
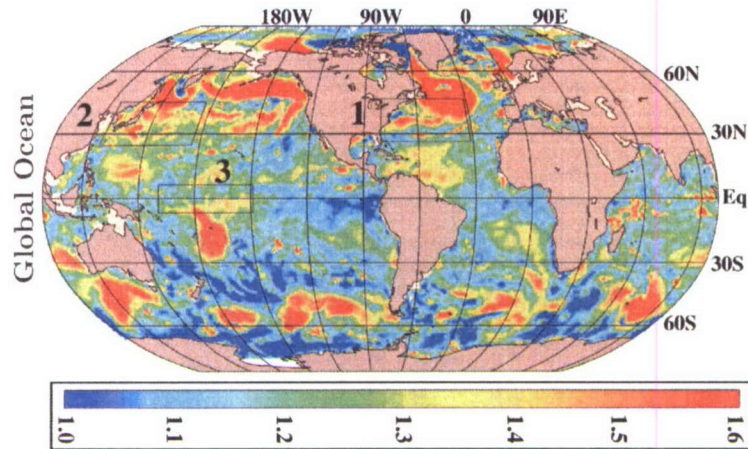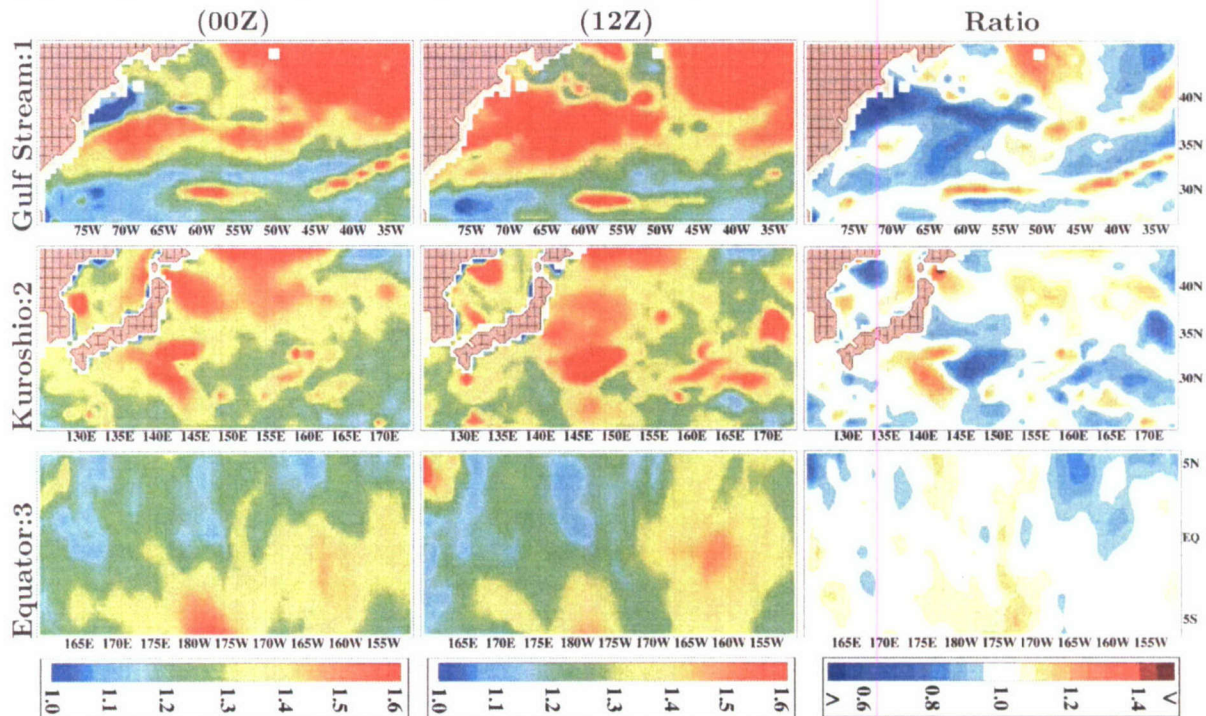
(a) Drag coefficient ($C_D \times 10^3$) over the global ocean on 6 Jan 2004 (00Z)

(b) Drag coefficient ($C_D \times 10^3$) at three zoom regions on 6 Jan 2004

FIG. 1. Values of drag coefficient (a) for the global ocean at 0000Z and (b) in 3 zoom regions numbered as 1, 2, and 3 (near the Kuroshio, Gulf Stream, and equatorial regions, respectively) at 0000Z and 1200Z. The ratio of drag coefficient [$C_D$(at 0000Z)/$C_D$(at 1200Z)] is also provided.

and waves is to reduce daily $C_D$ by <5% within those current systems.

## 4. Interannual variability of drag coefficient

Monthly mean $C_D$ fields, based on the method and data described (section 2), are shown in Fig. 2 over the global ocean from 1998 to 2004. For simplicity, we present mean $C_D$ fields in February, August, and November for each year. It is clear that there is no noise nor bull's-eyes in the fields, although no smoothing was applied to the $C_D$ fields. Note that in all panels, the regions where ice exists are masked out since $C_D$ over
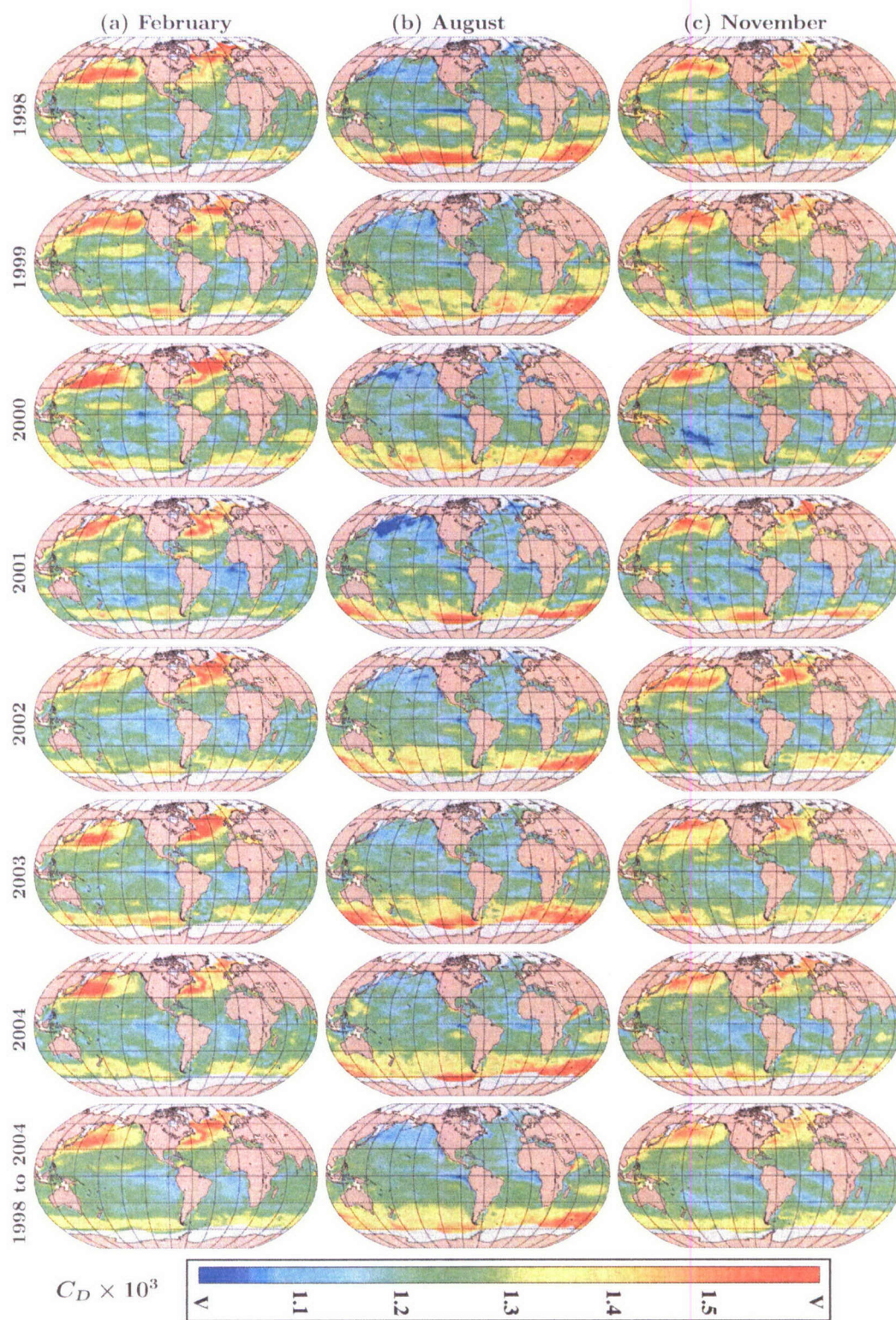
FIG. 2. Monthly mean drag coefficients over the global ocean in (a) February, (b) August, and (c) November. They are shown from 1998 to 2004, along with mean drag coefficients calculated over the 7-yr time period (bottom row). All values are in $10^3$. Regions where ice is present are shown in gray.

ice is quite different from that over the sea, and it is not the purpose of this paper to discuss sea ice drag coefficients. The ice-free regions over the global ocean (ice covered shown in gray) are determined using an ice–land mask based on data described in Reynolds et al. (2002).

The most significant feature evident from fields is that $C_D$ changes month by month during 1998–2004. This suggests that spatial and seasonal variations in $C_D$ need to be taken into account in climate applications, especially in the high northern and southern latitudes. On the contrary, there is little interannual variability for a given month. The $C_D$ values that are as large as $\approx \geq 1.5 \times 10^{-3}$ in February become as low as $\approx \leq 1.1 \times 10^{-3}$ in August over most of the North Pacific and North Atlantic Oceans, including the regions around the Kuroshio and Gulf Stream current systems. In fact, a close examination of spatial fields near the Kuroshio region reveals that there is an $\approx >50\%$ change or more in $C_D$ from February ($\geq 1.6 \times 10^{-3}$) to August ($\leq 1.1 \times 10^{-3}$). The $C_D$ value increases again in November. This is not surprising given the fact that there are even larger differences for a given day (see Fig. 1). The strong seasonality of $C_D$ is also evident at high southern latitudes.

Previously, Trenberth et al. (1989) also noted similar results, in that $C_D$ at high latitudes is generally larger than in the Tropics. However, $C_D$ values as low as $\leq 1 \times 10^{-3}$, which are seen in the eastern equatorial Pacific cold tongue in August and partially in November (Fig. 2), were not detected in their study. This is due mainly to the different gridded data and resolution used for calculating $C_D$ [the coarse-resolution $2.5° \times 2.5°$ ECMWF 1000-mb data (1980–86) versus the relatively fine $1.0° \times 1.0°$ NOGAPS surface data (1998–2004) used in this paper]. The winds at 1000 mb can also be very different from the winds 10 m above the sea surface. However, those were the only data available at that time. In addition, there were known deficiencies in the earlier ECMWF data, especially in tropical regions, as indicated in their study.

One important remark here addresses an aspect of uncertainty in the value of $C_D$ presented in Fig. 2. As mentioned previously, $C_D$ was calculated using atmospheric variables (air temperature and wind speed at 10 m, sea surface temperature, and relative humidity at the air–sea interface) from NOGAPS. Possible inaccuracies that may exist in these variables can be associated with errors in $C_D$. Thus, one might argue that there is not nearly enough information to make an estimate as to whether the errors in atmospheric variables from NOGAPS would outweigh any signal in $C_D$ variations discussed in this paper.

For the reasons mentioned above, we also calculate $C_D$ by substituting the similar atmospheric variables from another global dataset, the 40-yr ECMWF Re-Analysis (ERA-40) from the years 1979 to 2002 (Källberg et al. 2004). For simplicity, $C_D$ values calculated from NOGAPS and ECMWF data are compared for the months of January and June (Fig. 3). This is done for two different years, 2000 and 2001, to investigate the consistency of the $C_D$ comparison in different years. Clearly, there are no significant differences between $C_D$ values calculated from NOGAPS and ERA-40, as evidenced by $C_D$ ratios close to 1 over most of the global ocean (Fig. 3). The seasonal variability of $C_D$ is properly represented using either one of the datasets. Comparisons of atmospheric variables (e.g., wind speed, air temperature, etc.) between NOGAPS and ERA-40 are beyond the scope of this paper. However, as demonstrated here, both datasets give almost identical $C_D$ values over most of the global ocean, implying close agreement in surface atmospheric variables between the two.

Finally, we also investigate the impact of stability on $C_D$. It is possible that a user would like to compute $C_D$ based on only wind speed because air–sea temperatures and humidity may not be available or missing in an in situ dataset. Thus, the user would like to know the accuracy penalty when only wind speed is used for calculating $C_D$. To answer this question, more calculations are performed. Essentially, we use 6-hourly atmospheric variables from ERA-40, spanning 1979 through 2002. We calculated $C_D$ using (i) the air–sea temperature difference of zero along with relative humidity of 100 (i.e., neutral conditions) and (ii) the actual air–sea temperature difference and relative humidity (full stability). Both (i) and (ii) are done at each grid over the global ocean.

We first calculate $C_D$ and wind stress at each 6-hourly time interval and then form monthly mean $C_D$ and wind stress for each year. Finally, a monthly mean climatology is formed for both quantities over 1979–2002. This is done to provide an easier presentation of the results. The difference (long-term mean bias) between $C_D$ calculated based on full stability (ii) and that calculated based on neutral conditions (i) clearly demonstrates a significant effect of air–sea stability on $C_D$ even on monthly time scales (Fig. 4a). The $C_D$ including effects of the stability is generally larger over most of the global ocean. If one computed $C_D$ based on neutral stability as opposed to the full stability, errors could be very large and might be larger than 20% especially in the tropical regions (Fig. 4b).

## 5. Summary and conclusions

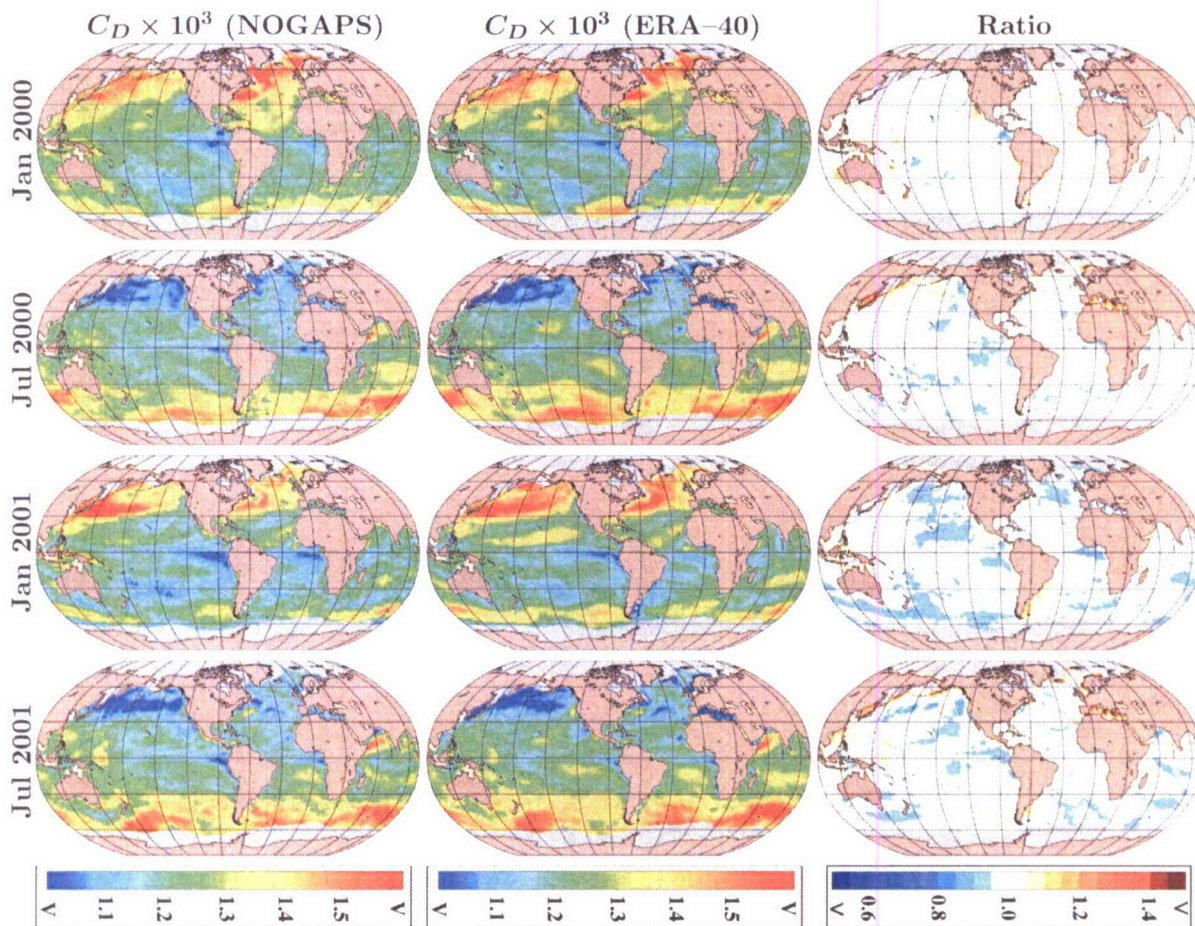There are no detailed studies examining the variability of $C_D$ on shorter (e.g., diurnal and daily) and longer

FIG. 3. Values of drag coefficient calculated using surface atmospheric variables from NOGAPS and ERA-40 as explained in the text. The ratio of drag coefficient [$C_D$(NOGAPS)/$C_D$(ERA−40)] is also provided and white represents ratios between 0.95 and 1.05.

(monthly, interannual, and climatological) time scales over the global ocean, even though there may exist field studies at specific time periods and certain places. This information about spatially and temporally varying $C_D$ over the global ocean is therefore long overdue in the literature. The aim of this paper is to fill in this gap. Such information is greatly desired by not only climate modelers but also modelers dealing with coupled atmosphere–ocean and ocean-only studies. In many cases, researchers dealing with such studies would like to see maps of these fields, as provided in this paper, to understand the variability of this particular quantity, especially for calculating wind stress.

In this paper, we use high temporal (e.g., 3-hourly) and fine spatial ($1.0° \times 1.0°$) resolution near-surface atmospheric variables from NOGAPS to determine stability-dependent $C_D$ on both interannual and climatological time scales. An examination of $C_D$ fields around the major oceanic current systems (e.g., the Kuroshio

and the Gulf Stream) demonstrates spatial variations not only for day versus night conditions but also for monthly time scales from 1998 to 2004. For example, a $C_D$ value of $1.0 \times 10^{-3}$ near the Gulf Stream at 0000Z can become as large as $1.8 \times 10^{-3}$ at 1200Z, implying a diurnal change ≈80%. This indicates the importance of including stability effects in calculating $C_D$, and thereby wind stress since air–sea temperature differences, relative humidity in addition to wind speed can be quite variable at those time periods. Outside the tropical regions, $C_D$ values are quite different from ≈$1.2 \times 10^{-3}$ and can even be ≥$1.6 \times 10^{-3}$ with strong seasonal variability—an important feature of $C_D$ that should be taken into account in air–sea interaction and climate modeling studies.

We also calculated monthly mean $C_D$ fields using 6-hourly atmospheric variables from the $1.125° \times 1.125°$ ERA-40 product. The $C_D$ values were found to be similar to those calculated from NOGAPS, indicat-

(a) Mean bias in "neutral" drag coefficient ($C_D \times 10^3$) (1979–2002)



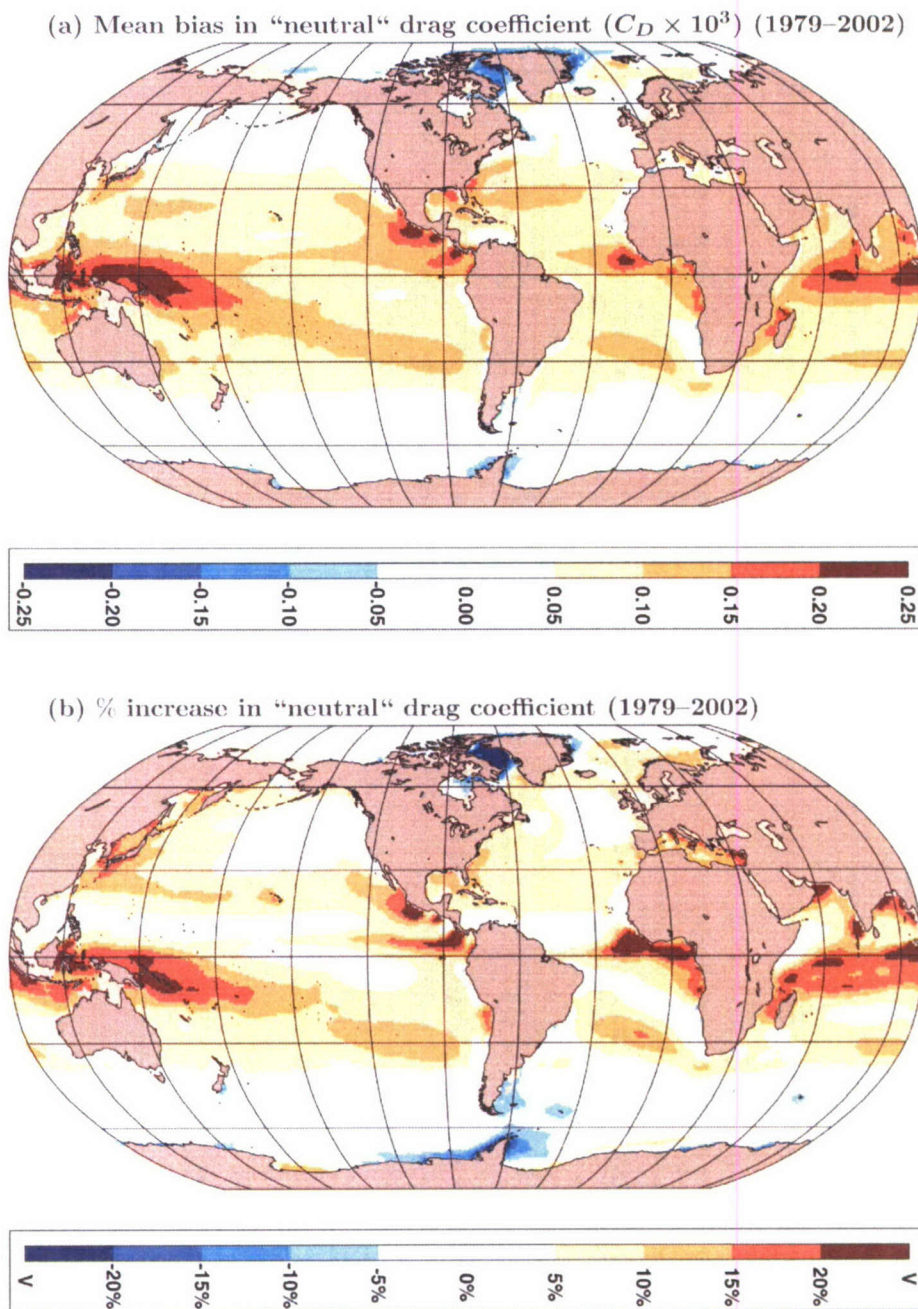(b) % increase in "neutral" drag coefficient (1979–2002)



FIG. 4. (a) Difference between the drag coefficient calculated using full air–sea stability dependence and the one calculated using neutral conditions. The latter is subtracted from the former. As explained in the text, differences in $C_D$ are calculated at each grid point over the global ocean during 1979–2002. (b) Percentage ratio of the drag coefficient computed using full air–sea stability to the one computed using neutral conditions over the global ocean during 1979–2002.

ing the robustness of the fields presented in this paper. One important conclusion arising from this paper is that ignoring the effects of air–sea stability in computing $C_D$ generally results in >10% error in comparison to the actual error that takes full stability into the account. Such errors are even larger (typically >20%) in the tropical regions. Finally, an examination of $C_D$ over the global ocean using other parameterizations may

yield slightly different results, but such an investigation is beyond the scope of this paper.

## REFERENCES

Bonekamp, H., G. J. Komen, A. Sterl, P. A. E. M. Janssen, P. K. Taylor, and M. J. Yelland, 2002: Statistical comparisons of observed and ECMWF-modeled open-ocean surface drag. *J. Phys. Oceanogr.*, **32**, 1010–1027.

Chelton, D. B., A. M. Mestas-Nuñez, and M. H. Freilich, 1990: Global wind stress and Sverdrup circulation from the *Seasat* scatterometer. *J. Phys. Oceanogr.*, **20**, 1175–1205.

Donelan, M. A., W. M. Drennan, and K. B. Katsaros, 1997: The air–sea momentum flux in conditions of wind sea and swell. *J. Phys. Oceanogr.*, **27**, 2087–2099.

Fairall, C. W., E. F. Bradley, J. E. Hare, A. A. Grachev, and J. B. Edson, 2003: Bulk parameterization of air–sea fluxes: Updates and verification for the COARE algorithm. *J. Climate*, **16**, 571–591.

Hwang, P. A., 2005: Temporal and spatial variation of the drag coefficient of a developing sea under steady wind-forcing. *J. Geophys. Res.*, **110**, C07024, doi:10.1029/2005JC002912.

Kållberg, P., A. Simmons, S. Uppala, and M. Fuentes, 2004: The ERA-40 archive. ERA-40 Project Report Series 17, ECMWF, 31 pp.

Kara, A. B., H. E. Hurlburt, and A. J. Wallcraft, 2005: Stability-dependent exchange coefficients for air–sea fluxes. *J. Atmos. Oceanic Technol.*, **22**, 1080–1094.

——, E. J. Metzger, and M. A. Bourassa, 2007: Ocean current and wave effects on wind stress drag coefficient over the global ocean. *Geophys. Res. Lett.*, **34**, L01604, doi:10.1029/2006GL027849.

Kessler, W. S., and R. Kleeman, 2000: Rectification of the Madden–Julian oscillation into the ENSO cycle. *J. Climate*, **13**, 3560–3575.

Koracin, D., C. E. Dorman, and E. P. Dever, 2004: Coastal perturbations of marine-layer winds, wind stress, and wind stress curl along California and Baja California in June 1999. *J. Phys. Oceanogr.*, **34**, 1152–1173.

Reynolds, R. W., N. A. Rayner, T. M. Smith, D. C. Stokes, and W. Wang, 2002: An improved in situ and satellite SST analysis for climate. *J. Climate*, **15**, 1609–1625.

Rosmond, T. E., J. Teixeira, M. Peng, T. F. Hogan, and R. Pauley, 2002: Navy Operational Global Atmospheric Prediction System (NOGAPS): Forcing for ocean models. *Oceanography*, **15**, 99–108.

Smith, S. D., 1988: Coefficients for sea surface wind stress, heat flux and wind profiles as a function of wind speed and temperature. *J. Geophys. Res.*, **93**, 467–472.

Sura, P., F. Lunkeit, and K. Fraedrich, 2000: Decadal variability in a simplified wind-driven ocean model. *J. Phys. Oceanogr.*, **30**, 1917–1930.

Taylor, P. K., and M. J. Yelland, 2001: The dependence of sea surface roughness on the height and steepness of the waves. *J. Phys. Oceanogr.*, **31**, 572–590.

Trenberth, K. E., W. G. Large, and J. G. Olson, 1989: The effective drag coefficient for evaluating wind stress over the oceans. *J. Climate*, **2**, 1507–1516.

Wuest, A., and A. Lorke, 2003: Small-scale hydrodynamics in lakes. *Annu. Rev. Fluid Mech.*, **35**, 373–412.